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Two climate-driven incision/aggradation rhythms in the middle Dnieper River basin, Central Russian Plain, in the Late Pleistocene

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Abstract

In valleys of River Seim and its tributaries (the middle Dnieper basin, west-central Russian Plain) two low terraces (T1, 10-16 m, and T0, 5-7 m above the river) and floodplain (2-4 m) with characteristic large and small palaeochannels exist. A complex of field and laboratory techniques was applied and three dozens of OSL and 14C dates were obtained to identify and establish chronology of incision and aggradation events that resulted in the current valley morphology. Two full incision/aggradation rhythms and one additional aggradation phase from the previous rhythm were recognized in the Late Pleistocene – Holocene climate cycle. The following events were detected. (1) Aggradation of Terrace T1 started in MIS 5 after the deep incision, probably in the end of MIS 6. (2) Incision into Terrace T1 occurred in late MIS 4 around 40 ka BP and probably reached elevation of the channel below the present-day river. (3) Aggradation of Terrace T0 started already in the end of MIS 3 and continued in MIS 2. Since ~25 ka and through the LGM lateral migrations occurred of a shallow braided channel located few meters above the present-day river. Channel and overbank alluvium of MIS 2 that makes the major part of terrace body is underlain by the late MIS 3 alluvium. (4) Incision into Terrace T0 started ~18 ka BP and proceeded below the modern river. Multiple-thread channel concentrated in a single flow, which at some places formed large meanders. In the period 15-13 ka BP, high floods that rose above the present-day floods left large levees and overbank loams on Terrace T0. (5) Aggradation to the modern position had occurred at the Younger Dryas – Holocene transition, so that small meanders of the Early Holocene already standed at the level of the modern river.

The result of river development in the late Pleistocene was the formation of three terraces, of which the two lower terraces – Terrace T0 and Floodplain – have very small difference in elevation and often merge. The main scarp in the bottom of the valley between Terrace T1 and Terrace T0/Floodplain was formed in the late MIS 4 incision phase. Incision phase in the end of MIS 2 (after the LGM) was undoubtedly governed by climatically forced large increase of water runoff. This phase consisted probably of two sub-phases at 18-16 and 15-13 ka BP separated by a relatively sub-phase of lower runoff. According to indirect reasoning, the late MIS 4 incision phase was also induced by considerable increase of water discharges. The established incision/aggradation rhythms are believed to be manifested over the Central Russian Plain outside the influence of ice sheets in the north and base level changes in the south.

Key words: fluvial geomorphology; river palaeohydrology; river terraces; palaeochannels; large meanders; aeolian processes; East Europe; Valdai epoch; Weichselian epoch

1. Introduction

1.1. River response to climatic variability in Western and Central Europe in the last glacial-interglacial cycle

River incision and aggradation has been extensively studied in Europe in the context of river response to environmental changes within the last glacial-interglacial climate cycle. Stratigraphic records from the cold part of the 100-ka thermal cycle sometimes reveal accelerating erosion activity and changing fluvial style in response to millennial-scale climatic oscillations such as Dansgaard–Oeschger and Heinrich events (Macklin et al., 2012), though this is not the rule and in many cases fluvial systems are sensitive only to long-term climatic changes within the Weichselian such as major climatic periods outlined by MIS boundaries (Mol, 1997; Kasse et al., 2003). A wealth of data has been obtained on fluvial dynamics since the Last Glacial Maximum (LGM) till the onset of the Holocene transition suggesting widespread occurrence of two common events – river aggradation around LGM and incision at the Pleistocene-Holocene transition, though details of their absolute chronology may vary in particular cases (Vandenberghe et al., 1994; Heine, 1999; Bridgland, 2000; Schirmer et al., 2005; van Balen et al., 2010; Erkens et al., 2011; Turner et al., 2013).

Much lesser degree of similarity is observed between fluvial reconstructions for the pre-LGM Late Pleistocene (Mol, 1997; Rose, Meng, 1999; Lewis et al., 2001; Kasse et al., 2003; Cordier et al., 2006; Antoine et al., 2007; Gábris, Nádor, 2007; Bridgland et al., 2008; Schulte et al., 2008; Brown et al., 2010; Schielein ewt al., 2011; Kaiser et al., 2012; Macklin et al., 2012; Nowaczinski et al., 2015). However consideration of a number of case studies over western and central Europe allowed Vandenberghe (1995, 2003) to synthesize that river incision phases correspond to climate transitions, both warm to cold and *vice versa*. Gibbard and Lewin (2002) summarized fluvial sedimentation patterns during the Middle and Late Pleistocene interglacials in lowland Britain and found the transitions from cold to warm climate stages to be dominated by lateral accretion and aggradation river modes and the major incision of rivers to have occurred in the beginning of subsequent glacial epochs. Similar conclusion has been reported from NE France where incision episodes coincided with warm-to-cold transition (Cordier et al., 2006). Bridgland and Westaway (2008) reviewed a large set of case studies worldwide and concluded that examples exist of both warming- and cooling-limb forcing of incision. Vandenberghe (2008) refined the model of river response to climate transitions in that the warm-to-cold transitions leave more distinct morphological evidence than cold-to-warm ones. However Cordier et al. (2014) found that not only cooling, but also warming transitions can play a major role in fluvial evolution.

Ambiguities in detection of climate signal may to some extent arise from that fluvial response to climate forcing in the Late Pleistocene was complicated in many European regions by local and regional tectonic patterns (Bridgland, 2000; Gábris, Nádor, 2007; Nádor et al., 2007; Westaway, Bridgland, 2014; Kiss et al., 2015), glacio-eustatic sea-level changes (Santisteban, Schulte, 2007; van den Schriek et al., 2007; Schulte et al., 2008), glacio-isostatic adjustment (Busschers et al., 2007, 2008; Bridgland et al., 2010; Panin et al., 2015), glacial and periglacial processes in alpine headwaters (Santisteban J.I., Schulte, 2007; Schielein et al., 2011; Rossato et al., 2013; Cordier et al., 2014), damming by ice sheets and release of glacial dams (Starkel et al., 2007; 2015; Panin et al., 2015). Centre of the Russian Plain drained by left tributaries of the middle Dnieper River is one of the most tectonically stable areas in Europe characterized by absence of long-term incision or aggradation trends due to crustal movements (Matoshko et al., 2002; Westaway, Bridgland, 2014). Remoteness from marginal sea basins and from Late Pleistocene ice sheet boundaries makes this region free from direct impact of glaciation, influence of glacio-isostatic movements and glacio-eustatic sea-level changes, and the fluvial history may be interpreted in terms of river response to climate change. At the same time, contemporary knowledge of alluvial stratigraphy in the central Russian Plain is still far from enough to provide a reliable climate-based explanation of incision/aggradation rhythmicity.

1.2. Late Pleistocene terraces in the central Russian Plain

Sandy alluvial terrace that rise by several meters above river floodplains are widely distributed in river valleys in the central Russian Plain and are called "first above-floodplain terrace", or "pine forest terrace". Typical elevation of this terrace above modern rivers is 9-15 m in the Oka River valley and its tributaries (Aseev, 1959; Kriger, Koposov, 1996; Rychagov, Antonov, 1996). In the Volga valley it rises from 8-11 m in the upper course to 17-20 m in the middle course (Moskvitin, 1958; Obedientova, 1977). In the Don valley, the "First Terrace" is recognized as the so called "Podkletnenskaya Terrace" with elevation 15-17 m, which aggradation is thought to have finalized in the beginning of MIS 2 (Grischenko, 1976). The "Second Terrace", which is associated with the first phase of the Late Pleistocene cold epoch, or MIS 5d – MIS 4, is recognized at the elevation 16-25 m (Grischenko, 1976; Kriger, Koposov, 1996; Rychagov, Antonov, 1996). Surface of this terrace is usually covered by loess-like deposits probably of aeolian origin, which is reflected in vegetation (not pine but mixed or broadleaved forests) and serves as a diagnostic criteria for differentiation of low terraces in river valleys.

In general, terrace spectra in the basins of Oka, middle Dnieper, upper and middle Don, middle Volga is rather uniform, which allows to correlate terraces in different valleys and to use similar geomorphic criteria for their designation and assigning of age. Valley development since the end of the Middle Pleistocene is usually schematized as succession of three incision-aggradation rhythms resulted in formation of three terraces: the Holocene floodplain, the First (8-12 m high) and the Second (15-18 m) Terraces associated with the two phases of the Valdaian (Weichselian) glaciation at MIS 2 and MIS 5d-4 respectively. Initial establishment of terrace chronology was based on tracing terraces and their passage into sandur plains associated with different ice-sheet limits (Moskvitin, 1958; Goretskiy, 1970; Obedientova, 1977). Accumulation of terraces during glacial epochs was confirmed by pollen analysis: pollen spectra of base parts of alluvial sequences evidence interglacial or interstadial conditions, and pass into cold-type spectra upward in alluvial sections (Grischenko, 1976; Antonov, Rychagov, 1996). According to pollen data, the upper parts of alluvial sequences represent only the first half of cold epochs, which allows to suggest that there were younger parts of cold epochs when river incision took place (Grichuk, Postolehko, 1982). On the other hand, a number of researchers proposed river incision to have started after the onset of the Holocene (Spiridonov, 1983; Butakov, 1986).

As seen from the above, river terrace formation in the central Russian Plain has still been studied mostly based on climatic stratigraphy and geomorphological correlation. It results partly from the lithological paradox, which is yet to be explained: though in Arctic and Subarctic regions – closest modern analogs of the Late Pleistocene environments – river floodplains are abundant with organic remains, organic finds in Late Pleistocene terraces in central Russian Plain are very rare. Scarceness of datable material predetermines rare employment of radiocarbon method, the most available tool for absolute dating. This is clearly seen from the fact that systematic 14C dates from river valleys in central Russian Plain begin from 13-14 ka BP (uncal) and refer mostly to floodplain sections (cf. Panin, Matlakhova, 2015).

In this paper, we present the first systematic OSL chronology of Late Pleistocene river terraces in this region. Field data collected in the valleys of River Seim and its tributaries complement our previous studies of floodplains of these rivers (Panin et al., 2001; Borisova et al., 2006). Taken together, previous radiocarbon and new OSL chronologies coupled with geomorphological evidence allow constructing a model of incision/aggradation rhythms – their time, position in the glacial-interglacial cycle and possible drivers.

2. Regional setting

River Seim is the left tributary of River Desna, which in turn tributes to middle Dnieper (Fig.1). The middle Seim fluvial system drains the south-west slope of the Srednerusskaya Upland. Total elevation range varies over the area between 100 – 120 m. Flat watersheds elevated at 230-260 m a.s.l. are subject to erosional dissection and acquire vertical relief of several tens of meters closer to trunk river valleys. Tops of the watersheds are cupped by Paleogene marine sands that serve as sources of sand fraction in river sediments. River valleys are incised into underlying Upper Cretaceous marine chalks, which leads to high clay and carbonate content in alluvia. Watersheds and higher river terraces are blanketed by 5-10-m thick Pleistocene loess cover. The maximal Dnieper glaciation correlated to the Saale II (Drente II) glacial phase in Europe (MIS 8) (Velichko et al., 2011; Gozhik et al., 2014) occupied the lower course of the Seim basin in the present-day territory of Ukraine. Maximal glacial boundary follows NNE to SSW 50 km west from the Svapa River, the westernmost area of our study (Fig. 1).

In the present-day landscape, the Seim valley makes a boundary in soil and vegetation cover: deciduous forests on grey forest soils (Retisols) spread to the north and forested steppes with Chernozems spread to the south of the valley. Modern climate is moderate continental with T_{JAN} and T_{JULY} -8º C and +19º C respectively and annual precipitation around 550 mm. About 70% of precipitation fall in the warm season, however both Seim and its tributaries are snow-fed rivers with high snowmelt flood in March-April (50-60% of annual runoff) and low water season, with rare and low rainfed floods, throughout the rest of year.

Valleys of Seim and Svapa Rivers are 80-100 m deep and contain four major terraces not counting the floodplain, of which the highest one (the Fourth Terrace) with elevation 50-60 m above rivers is assigned pre-Dnieper (before MIS 8) age (Milkov, 1983). Lower terraces make the 4-8 km wide alluvial plain at elevations 150-170 m a.s.l. This is a gently dipping surface composed of a number of merged terraces. The highest of them is the 20-25 m terrace (Second Terrace of Milkov, 1983). It can be correlated to the 20-25 m terrace of River Desna, which is thought to be of the Early Valdai (Early Weichselian, MIS 5d-4) age (Velichko et al., 1977). The lowermost terrace (not counting the floodplain), or First Terrace by Milkov (1983), can be subdivided into two steps 12-15 and 7-10 m above rivers. In the Lgov reach of Seim and in lower Svapa River this terrace is loess-covered and bears numerous flat-bottom depressions 20-30 to 80-100 m in diameter – padings, or "steppe dishes" in Russian terminology, interpreted as relic thermokarst sinks (Panin et al., 2001). At the Kurchatov reach of Seim this terrace is sand-covered with numerous relic aeolian dunes. Few sand massifs on this terrace can also be found in the Lgov – Svapa area (Panin et al., 2001).

Floodplains of Seim at Lgov and lower Svapa Rivers bear signatures of the lateral migration of large meandering flows such as large palaeomeanders and associated ridge-and-swale topography (Fig. 2). In the Seim valley, large palaeochannels have meander half-wavelength around 3000 m and bankful channel width 350 – 500 m, which is 5-10 times larger than that of the Holocene palaeochannels and present-day river. In the lower Svapa valley these parameters are 1400 m and 250-350 m respectively and similar ratios with the Holocene and modern values. In both valleys, 2-3 age generations of large palaeomeanders exist, which evidences rather long duration of the period of their formation. Based on radiocarbon dates from the base of their infills (Panin et al., 2001; Borisova et al., 2006), large palaeomeanders were formed after the LGM since not later than 14.0 uncal ka BP and till not earlier than 12.6 uncal ka BP. Paleofloristic data and modern analog approach were exploited to quantitatively estimate hydroclimatic parameters in the Seim River basin and it was found that large meander formation was forced by high river discharges produced by mean annual runoff depth three times as large as today (Borisova et al., 2006). Holocene palaeochannels have parameters similar to that of the modern river. Characteristics of the Seim and Svapa valleys at study sites are presented in Table 1.

Previous research at the study sites (Panin et al., 2001; Borisova et al., 2006) were focused on river planform changes – formation of large palaeochannels and their transformation into small meanders in the Late Glacial – Holocene time. These studies were limited to river floodplains and left aside river terraces because they do not bear preserved traces of river migration and their chronology could not be established as terrace sands do not contain materials datable by radiocarbon (no organic remains, ages outside the range acceptable by radiocarbon technique). Geological data was based on relatively shallow hand cores that almost never reached underlying rocks. To establish the Late Pleistocene incison/agradation rhythms, estimate their amplitude and geochronology, new field methods and laboratory techniques had to be used.

3. Methods

3.1. Field techniques

Accurate determination of landform elevation above the river was conducted at study sites by DGPS topographic profiling performed with Leica Smart Station 1200. Differential correction into GPS measurement was made partly in real-time regime with radio-modem, partly in post-processing mode. Final accuracy of topographic measurements was at least 10 cm. Ellipsoid WGS84 heights were converted to orthometric heights, i.e. elevations above sea level used on topographic maps, with the EGM2008 geoid model.

Sedimentary composition of river terraces was studied in river-bank exposures and pits, in mechanical cores (up to 20 m deep) and by means of Ground Penetration Radar (GPR) technique (GPR Zond 12e with shielded antenna 500 mHz). Sampling for OSL analysis was performed from exposures or cores into opaque metallic or plastic tubes. For grain size analysis bulk samples were collected from lithologically uniform units of studied sections. Samples were also taken from present-day river bed and point bar as a reference for assessment of sedimentary facies. Sampling for pollen analysis was conducted from 1 cm units at 5-10 cm interval from fine-grained lithological divisions. In sandy divisions sampling spacing was uneven. Given that pollen concentration in sand deposits is usually low, we tried to take samples from more loamy parts and avoided pure sands if it was possible. Such sampling style is grounded on relatively high vertical aggradation rates characteristic for alluvial deposits, so that spacing of samples does not influence much the interpretation of a pollen diagram.

3.2. Laboratory analyses

Samples for grain size analysis were dried at temperature 95^o, mixed and weighted. Analysis was performed for 30 g sub-samples. Fractions larger than 0.05 mm were separated by dry sieving with Fritsch Analysette 3 PRO vibratory sieve shaker. Fractions finer than 0.05 mm were analyzed with the Fritsch Analysette 22 laser particle sizer.

Samples for pollen analysis were prepared using the pollen extraction procedure developed by Grichuk (1940). The treatment included separation by heavy liquid (cadmium iodide) with a density of 2.2 g/cm3. Calculation of relative pollen frequency is based on the total terrestrial pollen sum including arboreal pollen (AP) plus non-arboreal pollen (NAP) plus spores. Aquatic plants were excluded. As pollen content in studied sediments was low, all available pollen and spores were counted in samples. Morphological determinations of pollen were carried out according to Reille (1992). For calculating pollen concentrations, Lycopodium tablets were added to each sample during the pollen preparation process (Stockmarr 1971). Pollen diagrams were constructed using Tilia and Tilia Graph programs (Grimm 1990).

Microscopic study of quartz grains was performed to facilitate designation of sedimentary facies, primarily the differentiation of alluvial and aeolian sands. In the study area, aeolian sands origin from reworking of alluvial deposits. The duration of wind action may have been relatively short, which results in weakness of aeolian signals on grains surfaces and therefore in difficulties in identification of sand reworking by aeolian processes. This limits the effectiveness of using optical microscopic studies and makes necessary to apply Scanning Electron Microscope (SEM). We used SEM JEOL JSM-661 LV at magnification from 160× to 400× for whole grains and up to 1800× for selected parts of grains. Features typical for alluvial and aeolian origin were identified according to published sources (Krinsley, Doornkamp, 2011; Bull, Morgan, 2013). Aeolian sediments are characterized by well- or perfectly-rounded outline and matt surface with numerous small pits – so-called upturned (cleavage) plates (Fig. 3A). Alluvial sediments are characterized by average- or well-rounded outline, smooth glossy surfaces and V-shaped irregularly oriented impact pits (Fig. 3B). Eight samples from the Seim valley at Kurchatov site were analyzed (Table 2). In each case an aliquot of 25 grains from fraction 0.25-0.5 mm was used. Origin of each sample was assessed on the basis of examination of SEM images and counting grains with specific fluvial and aeolian features in each aliquot.

Alluvial grains reworked and non-reworked by aeolian processes were found to have many common characteristics, however a number of specific features was identified to separate between them (Matlakhova, 2015). Aeolian features may sometimes overlay over fluvial features, which makes the evidence of wind reworking of initial alluvial sediments. Nevertheless, such clear evidencing is not a common case. One of the main differences of the grains reworked by aeolian processes is presence of upturned (cleavage) plates on their surfaces, but they occupy less area than on "classical" aeolian grains and are present preferably on corners of grains. Also the aquatic features, first of all, V-shaped irregularly oriented impact pits and smooth glossy surfaces, occupy less areas on reworked grains than on alluvial grains and preserve only in selected areas of grains, mostly in depressions on surface which are better protected against aeolian action. Also alluvial sediments reworked by wind are slightly better rounded than non-reworked alluvial grains, but less rounded then "classical" aeolian grains. Usually "pure" alluvial grains have less dissolution-precipitation features than reworked alluvial grains. Also the number of some other aquatic features (large and small conchoidal fractures, curved grooves, et. al.) is less on reworked grains. Thus samples that we marked as aeolian (reworked alluvium) deposits are composed predominantly of grains bearing aeolian features; usually more than 90% of grain area have clearly expressed signatures of aeolian action, firstly on the corners, but also sometimes all over the grain surfaces.

3.3. Dating techniques

Geochronology is based mainly on OSL dating produced in two laboratories – Nordic Laboratory for Luminescence Dating (NLL, Aarhus University, Denmark; lab index Risø) and GADAM Centre of Excellence (Silesian University of Technology, Poland; lab index GdTL) (Table 3). In the field, samples were collected using opaque plastic tubes from vertical cleaned sections or from cores.

Sample treatment for dose rate analysis in the GADAM Centre started with drying and storage for ca. 3 weeks prior to measurement to ensure equilibrium between gaseous ^{222}Rn and ^{226}Ra in the ^{238}U decay chain. For dose rate determination, high-resolution gamma spectrometry with Canberra HPGe detector was carried out in order to determine the content of U, Th and K. Each measurement lasted for at least 24 hours. The activities of the isotopes present in the sediment were determined against IAEA standards RGU, RGTh, RGK after subtraction of the detector background. At the NLL dating, the outer tube material (~200 g) was dried, ashed for 24h at 450°C, ground to <200 µm and then cast in wax cup to provide a constant counting geometry. After three weeks these were counted for 24 h on a laboratory gamma spectrometer as described in Murray et al. (1987). Dose rates were calculated using the conversion factors of Adamiec and Aitken (1998). For the NLL samples a 20±10% Rn loss compared to its parent ^{226}Ra is assumed. The cosmic ray dose-rate contribution at the sites was estimated as described by Prescott and Hutton (1994). Average lifetime depth of samples relevant for this estimation was assigned with respect to erosion/sedimentation history of each site.

Equivalent doses were determined using the single-aliquot regenerative-dose (SAR) protocol (Murray and Wintle, 2000). Coarse grains of quartz (90–125 µm) were extracted from the inner tube sediment by standard laboratory practice. The sand-sized fraction was treated with 20% hydrochloric acid (HCl) and 20% hydrogen peroxide (H_2O_2) first. Quartz grains were then separated using density separation using of sodium polytungstate solutions leaving grains of densities between 2.62 g/cm³ and 2.75 g/cm³. This fraction was treated with concentrated hydrofluoric acid (HF) to remove remaining feldspar grains and etch away the outer alpha irradiated rind of the quartz grains. After etching the grains were sieved again. OSL measurements were performed using an automated Daybreak 2200 TL/OSL reader on 6 mm large aliquots. Blue light stimulation was carried out by the in-built array of blue LEDs (470±4 nm) delivering about 60 mW/cm² at sample position. Laboratory irradiations were made using a calibrated $^{90}Sr/^{90}Y$ beta source integrated to the reader delivering a dose rate of 5.27 Gy/min. Similar standard measurement and sample preparation procedures were undertaken at the NLL except for the 2.75 g/cm³ density separation and the resieving step after HF etching. The purity of the quartz extracts mounted as large (8 mm) aliquots was confirmed by the absence of a significant infrared stimulated luminescence (IRSL) signal and clean 110°C quartz TL peaks. All SAR measurements were undertaken using Risø TL/OSL DA-20 readers with the sample held at 125°C during 40 s of blue light stimulation (470 nm, ~80 mW/cm^2). Visual inspection of the decay curves showed that all the samples are dominated by a strong fast component. Preheat conditions for the natural/regenerative dose signals were 260°C (duration 10 s) and 220°C (duration 0 s) for the test dose signals. After each SAR cycle the aliquot was stimulated with blue light for 40 s at elevated temperature (280°C). Using these parameters a dose recovery test was carried out on 3 to 6 aliquots of each sample. After bleaching the aliquots with blue light at room temperature (two 40 s stimulations with a 10 ks pause in between) they were given a beta dose close to the measured ED value. The average ratio of the measured to given dose is 0.99±0.02 (n=42) suggesting that the chosen SAR protocol is suitable for dose measurement. For the sake of comparability with calibrated radiocarbon ages, OSL ages were calculated before AD 1950.

New and unpublished radiocarbon dates complemented with a number of previously published dates relevant for this study are presented in Table 4. Dating was performed in Kiev radiocarbon laboratory (Ukraine; lab index Ki), Geological Institute of Russian Academy of Science (index GIN) and GADAM Centre of Excellence (Poland; index GdA). All dates but the one AMS date (GdA) were produced using liquid scintillation technique and calculated as conventional radiocarbon ages. To achieve comparability with OSL ages, radiocarbon ages in the text and in figures are presented as calendar years in BP (prior to AD 1950) format. Calibration was performed in the OxCal 4.2 program (Bronk Ramsey, 2009) using IntCal13 calibration curve (Reimer et al., 2013).

4. Results

4.1. Seim valley at Kurchatov

Study site is located at the Kurchatov town 25 km downstream from the Kursk city (Fig. 1). The valley has total width of some 10 km, of which 3-4 km is occupied by floodplain and lower terraces with elevations up to 16 m above the river (Fig. 4). Field topographic survey (DGPS profiling) allowed to recognize a number of terrace steps combined into three terrace levels (Fig. 5). Holocene floodplain with high-amplitude levee-hollow morphology and dominant elevation 3-4 m above the river. Floodplain is abundant with Late to Mid-Holocene oxbow lakes that evidence lateral movements of a meandering channel similar to that of the present-day river. Terrace T0 (5-7 m) occupying the most part of the valley bottom has smooth topography bearing no traces of lateral migration of the river channel (Fig. 4, 5). At many places it has no distinct boundaries and merges with the floodplain. Terrace T1 (12-16 m) can be associated with the "first terrace" by Milkov (1983).

4.1.1. Terrace T1, Section ML1

Composition of terrace T1 was studied in detail in an old sand pit at the Maliutino village. A 500 m long GPR profile along the western wall of an old sand pit cut into terrace T1 revealed three distinct radar facies (Fig. 6). The main body is terrace alluvia with a number of undulating reflection boundaries within it and no clear reflection pattern below 11-12 m due to appearance of ground waters. The upper 2-3 m thick layer with clear lower reflection surface and horizontal bedding within it was interpreted as aeolian cover sands, which was confirmed by further lithological study of section ML-1 (see below). At the distance 190-250 m, a 7-8 m deep trough was detected with a wedge-like lower part and widening top that abuts at the bottom of the aeolian cover. In the field, we interpreted the trough to be a palaeochannel and decided to clean the pit wall at its side so that to study both palaeochannel fill and underlying terrace alluvia, however this interpretation was re-assessed after examination of the cleaned section (see below).

The upper 10 m of the fill of Terrace T1 were examined in section ML1 (51.67690ºN, 35.77565ºE, 172.8 m a.s.l.; 12.0 m above the river). The section was divided into 7 lithostratigraphic units (from top to bottom; Fig. 7A).

Unit ML1-1, 0.0 – 2.4 m: well-sorted light-yellow fine sand with alternating massive structure and thin horizontal bedding, with ferruginous layers (ortsands) from top to the depth of 1.3 m. At depth 1.7 m an 1.5 cm thick horizontal layer of tight ferruginous loamy sand was found underlain by compacted silty sand (rudimentary soil?). With this layer a small wedge cast is associated interpreted as frost crack. In the upper 0.7-1.0 meter the layer is disturbed by shallow anthropogenic pits filled with mixed charcoal-rich sand. The lower 30 cm of the unit are composed of loamy sand with indistinct cryogenic deformations such as involutions, microfolds and boiling pots. The lower boundary of the unit is sharp and wavy, probably due to cryogenic action.

Unit ML1-2, 2.4 – 3.3 m: very tight unevenly colored (reddish-brown spots on greenish-grey background) sandy loam heavily disturbed by cryoturbations; intrudes into the underlying sands in the form of microdykes (signatures of frost cracks?).

Unit ML1-3, 3.3 – 3.7 m: brownish fine sand with thin horizontal parallel to slightly wavy lamination, with few particle-thick lenses of coarse sand (aeolian micro-pavement?), with clear and sharp (by color and grainsize) lower boundary.

Unit ML1-4, 3.7 – 5.0 m: interbedding (beds 5-10 cm thick) of greenish-grey loam containing 1-2 cm thick lenses of fine sand, and fine sand with reddish brown ferruginous spots and beds. The whole unit is cryoturbated. Below 4.4 m beds become thinner (sand 1-3 cm, loam 0.5-1.5 cm) and acquire dipping to the east (upstream the river). Individual beds are concave with slope changing downwards from 30º to 15-20º. In the right wall of the section clinoform bedding is cut by a large wedge cast filled with light reddish-brown fine sand; the right part of which is cut by terrace slope. At the contact with the wedge, alluvial beds are bended up, which indicates lateral growth of former ice wedge. The wedge goes down from the upper part of Unit 4 to depth 5.0 m and has characteristic "shoulders" dividing wide upper and narrow lower parts at depth 4.4 m (former contact of permafrost and active layer?). In the lower part the half-width of the wedge preserved in the section is 20 cm wide. The whole Unit has an erosional lower contact that dips in the upstream direction.

Unit ML1-5, 5.0 – 6.2 m: greyish fine sand with 1-2 cm thick horizontal layers of dark-brown loam repeated in a 20 cm interval. The upper loam layer at depth 5.1 m shows load structures in the form of potato-like bulges hanging down from the layer and composed of the same loams. The ferruginous lower 20 cm of the unit are highly cryoturbated.

Unit ML1-6, 6.2 – 9.6 m: grayish with brownish layers, horizontally bedded fine sand with layers and lenses of medium to coarse sand. Highly deformed at the top 10-15 cm by cryogenic action. Clear erosion contact with the underlying unit.

Unit ML1-7, 9.6 – 10.1 m (visible): whitish fine sand, predominantly quartz, horizontally bedded, with brownish layers of loamy sand.

SEM analysis was performed for samples from five units of the section (Table 2). Its results allowed to propose the following sedimentation mechanisms: Units 2, 5 and 7 – fluvial, Units 1 and 3 – aeolian (relatively short wind transport of initially alluvial grains).

Facial interpretation of the recognized units allowed to join the lithostratigraphic units into four groups, each of which were OSL dated (Table 3, Fig.7A). The lowermost Unit 7 was interpreted as the oldest alluvial sediments in the section and dated to 77±7 ka BP. Units 6 and 5 make the second alluvial package representing active channel facies and facies of low floodplain respectively. Unit 1-5 gave OSL age of 50.1±3.9 ka. The next package consisting of Units 4 – 2 is the fill of the negative form detected in the GPR record (Fig. 6). Of the three units, only the lowermost Unit 4 with OSL age 17.8 ±1.0 ka can be associated with alluvial sedimentation. Unit 3 is regarded to have been accumulated by aeolian processes. The uppermost Unit 2 dated at 15.9 ± 0.8 ka, which combines massive texture and relatively thin grain size, looks as the result of very slow accumulation of products of sheet erosion – slopewash, or deluvial, deposits. The last group consists of the topmost Unit 1 – aeolian sand cover. OSL date 21.7±1.6 ka derived from this deposit makes the inversion in relation to the two other dates below in the section.

4.1.2. Terrace T0, Section ML2

Section ML2 (51.68062ºN, 35.77914ºE, 167.2 m a.s.l.; 6.4 m above the river) was found and first studied by A.L.Chepalyga (2000). It is ~100 m long cliff undercut in the outer bank of a steep meander oriented normally to the valley axis from Terrace T1 to the center of the valley (Fig. 8). Our examination of the exposure combined with coring to the depth of 11 m allowed to recognize five lithostratigraphic units in the section (Fig. 7B).

Unit ML2-1, 0.0 – 1.4 m: tight heavy loam with massive structure, dark grey to black in the top (chernozem soil) gradually changing downwards to dark-brown. Unit thins out to the left side 150 m from the terrace edge at elevation and ~7 m above the river. To the right, the unit thickens up to 1.6 m. It continues on the terrace scarp under the floodplain deposits leaning the terrace and descends down under the river (Fig. 8). At the terrace edge, the unit is destroyed by artificial earthworks.

Unit ML2-2, 1.4 – 2.6 m: light yellow medium sand, brownish and cemented in the top 25 cm (rudimentary soil?) and loose throughout downwards, with clear horizontal microlamination to the depth 190 cm and massive structure below, with lenses of coarse sand (most thick is the one at depth 186-190 cm: aeolian pavement?) with sharp bottom contact.

Unit ML2-3, 2.6 – 4.8 m: fining downwards horizontally bedded sequence starting from ferruginous medium sand (typical beds 1 cm thick) in the upper half-meter and finishing with the 45-cm thick layer of grayish-brawnish microlaminated heavy loam in the bottom extending along almost the whole exposure (Fig. 8).

Unit ML2-4, 4.8 – 7.7 m: sequence similar to the previous unit; the layer of heavy loam in the base is ~1 m below water in the left side of the exposure and rises to the water surface in the right (Fig. 8).

Unit ML2-5, 7.7 – 11.0 m (visible): watery mixed-size sand, probably with silty beds.

SEM analysis conducted for two samples from the section allowed to identify Unit 2 as having aeolian origin and Unit 4 as fluvial (Table 2).

Facial interpretation and geochronology of the section are as follows. Unit 5 is channel alluvia, which sedimentation style can not be reconstructed because of liquefied state of material in the core. Units 4 and 3 are channel alluvia deposited in a river with contrast bottom topography. Base layers of loam were accumulated in pools that were subsequently buried by shifting side- or mid-channel bars. Age of the lower Unit 4 is supported by OSL dates 33.1±3.0 ka and 31.8±1.6 ka and 14C date 27400±580 BP, or 31590±620 cal BP. Unit 3 was OSL dated at 25.3±1.2 ka BP. Unit 2 is aeolian sand that descends from the nearby sand terrace. It gave an OSL date of 19.1±1.1 ka. The upper 25 cm were compacted and cemented due to long exposure and may be regarded as rudimentary soil. The top of this pedogenic sub-unit gave a radiocarbon (AMS) date of 8700±25 BP, or 9630±50 cal BP. However, the same soil at the terrace scarp covered by loams of Unit 1 was dated to 12600±150 BP, or 14840±310 cal BP (Fig. 8). Unit 1 is overbank alluvium. It gave two contradicting OSL dates: 9.9±0.8 ka and 13.6±0.7 ka.

Because of high degree of bioturbation of Unit ML1, samples for pollen analysis from the uppermost Unit ML2-1 were taken from an individual column few meters apart from the main section – subsection 2 on the pollen diagram that overlaps with subsection 1 corresponding to Units ML2-21 – ML2-4 (Fig. 9). Four pollen assemblage zones (PZ) were determined in both subsections.

PZ ML2-1, depth 700-140 cm. The pollen content in sediments lying below 140 cm is extremely low. Only rare pollen grains of *Betula*, *Pinus*, *Asteraceae*, *Poaceae*, *Cyperaceae* and spores of *Polypodiaceae* were registered. These taxa are characterized by wide ecological requirements and occurred often in pollen assemblages from sediments formed in glacial epochs (Zelikson, 1995; Novenko, 2006). The low pollen concentration can be explained by lithology of deposits represented mainly by sand and low pollen productivity of vegetation during the Last

Glaciations. An exception is the sample from depth 520 cm, just bellow the OSL data 33.1±3.0 ka BP, in which pollen concentration slightly increased. The pollen assemblage is dominated by arboreal pollen (60%), represented by *Pinus* and *Betula*, including tree and shrub species of birch. *Picea* pollen occurs in small quantity. It should be noted that spruce pollen was found in the depth interval 600-500 cm but it is absent in sediments at the depth 500-150 cm. Among herbs *Asteraceae* and *Poaceae* pollen was identified, spores of *Equesetum* and *Polypodiaceae* were recorded.

PZ ML2-2, depth 140-90 cm. AP values in pollen assemblages are relatively high (40-55%). The total pollen concentration rises toward the top of this zone. Pollen spectra are dominated by *Pinus* and *Betula*, pollen of *Picea* is less abundant. Pollen of broadleaf trees (*Ulmus* and *Quercus*), *Corylus* and *Alnus* was found in this pollen zone. Pollen of herbaceous plans is relatively diverse and represented mainly by *Poaceae*, *Artemisia*, *Cichoreaceae*, *Asteraceae* and *Polygonaceae*. Pollen value of *Polygonum* aviculare-type, species typical for disturbed grounds, reaches up to 20% in pollen assemblage of sample from subsection 2. Group of spores includes *Polypodiaceae*, *Equesetum* and *Botrychium*.

The pollen concentration of sand deposits in the middle part of the subsection 2 (depth 140-45 cm) is very low making it impossible to assess the proportion of components in pollen assemblages. The upper part of the section was divided into two pollen zones.

PZ ML2-3 (45-25 cm). Pollen assemblages are characterized by the highest content of AP (up to 80%). *Betula* pollen achieves maximum (60-70%), *Pinus* pollen comprises not more than 20%. Pollen of other trees and shrubs (*Quercus*, *Ulmus*, *Tilia*, *Alnus* and *Corylus*) is poorly represented. The NAP includes *Poaceae, Artemisia*, *Cichoreaceae*, *Rosaceae*, *Ranunculaceae*, *Asteraceae*, *Apiaceae*, *Caryophyllaceae*, *Fabaceae*, etc. Among spores that of *Ophyoglossum vulgatum*, *Sphagnum* and *Athoceros* appeared.

Pollen spectra of Zone ML2-4 (25-0 cm) are characterized by an increase in NAP content (70-75%) and a high diversity of the herbaceous plants. *Poaceae*, *Chenopodiaceae*, *Ateraceae* and *Cichoreaceae* dominate in this group. Pollen of species – anthropogenic indicators such as cultivated cereals, *Plantago*, *Rumex*, *Fagopyrum*, spores of *Athoceros* and *Riccia* are frequent.

The terrace is adjoined by the Holocene floodplain with 14C date of 2160±90 BP (2160±110 cal BP) obtained from channel alluvium (Fig. 8). Humus from buried soil on the surface of terrace scarp buried under the floodplain overbank alluvium was dated at 3530±150 BP (3840±200 cal BP). Artificially mixed loams at the terrace-floodplain contact were dated to 680±140 BP (660±120 cal BP), which may be associated with the period of human disturbance of the terrace edge.

4.1.3. Terrace T0, Section ML3

Section ML-3 (51.67514ºN, 35.78486ºE, 166.7 m a.s.l.; 5.9 m above the river) is the exposure of the 6 m terrace at the right bank of Seim 700 m upstream from Section ML2, which was continued 3.5 m below the river with a hand core (Fig. 4). Terrace surface bears a gentle ~0.5 m high levee oriented in the W-E direction, i.e. normal to the valley. The described and sampled section is on the southern (upstream) slope of the levee near its top. Five lithostratigraphic units were identified in the section.

Unit ML3-1, 0.0 – 0.45 m: dark grayish-brownish humus rich heavy loam, with admixture of fine sand, with silicate powder below 0.35 m (eluvial soil horizon?). 1-3 cm large fragments of ceramics were found at depth 10-1 and 23-24 cm. Ceramics was dated to Early Iron Age (expertize made by G.Starodubtsev, director of the Kursk Regional Archaeological Museum).

Unit ML3-2, 0.45 – 1.35 m: tight brown heavy loam with no lamination, reworked by soil processes (clayey and ferruginous cutans: illuvial soil horizon?)

Unit ML3-3, 1.35 – 4.65 m: interbedding of greyish-brownish silty loam (beds 0.5-1 cm thick) and light-yellow silt (1-2 cm beds). In the upper part, beds are horizontal. At depth 2.15-2.20 m a slightly darker color may evidence rudimentary soil formation and thus short interruprion of sedimentation. In the interval 2.2-4.1 m carbonate concretions are found up to 2 cm large. Down from 2.2 m beds are dipping to the south (upstream) and to the west (riverward) at angles 10-20°, conformal to the levee seen in the terrace surface (see above). Down from depth 3.5 m, substantial admixture of fine sand appears within the silt beds. The sharp erosional bottom boundary of the unit dips upstream (from 4.65 to 4.75 m within the cleaned section 1 m wide) and is marked by laminae of fine sand dipping upstream (to the right).

Unit ML3-4, 4.65 – 8.90 m: light grey, bluish below 5.65 m heavy loam with microlamination that emerges on washed surfaces below the water level; silt and fine sand content increases down from 5.30 m.

Unit ML3-5, 8.90 – 9.70 m (visible): silty fine to medium sand.

Grain size and lamination style allowed to interpret the lowermost Unit ML3-5 as channel alluvium and Units 1 to 5 as horizons of overbank alluvia accumulated during different periods at and after the LGM, except for the lower pat of Unit ML3-3 (3.5 – 4.65 m), which may be considered as low energy channel alluvia (Fig. 7C). Unit ML3-5, both the channel and overbank alluvial facies were OSL dated to similar ages around 19-20 ka BP: bottom of the Unit gave the date 19.8±1.3 ka, top of the Unit – dates 19.1±1.0 ka and 18.5±1.0. Unit ML3-2 gave two OSL dates from the Late Glacial: 14.1±0.9 ka and 15.4±0.7 ka. The uppermost Unit ML3-1 accumulated in the Late Holocene, based on the ceramics finds.

Pollen analysis was performed for the entire section; nevertheless, only pollen records from the lowermost and uppermost parts allowed the reliable reconstructions of vegetation (Fig. 10). Sediments at the depth 400-150 cm contain almost no pollen: only rare grains of pine and birch were found. Pollen content in intervals 500-400 cm and 150-55 is very low to calculate percentages of the main components. Only three pollen assemblage zones were identified in the section, as follows.

PZ ML3-1 (560-465 cm) corresponds to the layer of clayey silt in the base of the section. Pollen assemblages are dominated by AP (80-95%), mainly by *Pinus* and *Betula*, meanwhile pollen values of dwarf birch do not exceed 5%. Pollen contents of Picea reach 20-35%. Among NAP *Poaceae*, *Polygonaceae*, *Cyperaceae* are relatively abundant. The presence of *Nymphaea* pollen, that is rather thermophilous plants and not typical for the glacial epoch, indicates some warming.

PZ ML3-2 (55-47 cm, the OSL data 15.4 ±0.7 ka BP at the depth 50 cm). Pollen spectra are marked by an increase of AP content (80-95%), mainly by pollen of pine, spruce and birch, both tree and shrub species. *Picea* forms a local peak in this zone (25-30%), and *Pinus* achieves maximum (up to 70%). NAP group is poor and represented by pollen of *Cyperaceae* and aquatic plants – *Typha latifolia* and *Nymphaea*.

PZ ML3-3 (47-0 cm). Pollen assemblage of the lower part of the zone are characterized by maximum of *Betula* pollen that followed by increase of broadleaf tree pollen values. In the upper part of the pollen zone AP values decreased while herbaceous pollen such as *Poaceae*, *Chenopodiaceae*, *Polygonaceae*, *Rosaceae*, *Asteraceae* and *Cichoreaceae* become plentiful. A permanent component of pollen spectra is spores of ferns.

4.1.4. Avdeevo site

In the NE side of the valley bottom, a well-known Late Palaeolithic archaeological site is located on a remnant of terrace T0 (Fig. 4). According to the DGPS profile, the present-day terrace surface at the site is elevated by 6.5-7.0 m above River Seim (Fig. 5). Cultural layer with Eastern Gravettian finds (Gvozdover, 1995) lies 1.2 – 1.5 m below the surface (Fig. 11A). Modern flood stage in the Seim valley does not exceed 5 m. Nevertheless, the Avdeevo site is located at the bank of the Rogozna River 2 km upstream from its confluence with Seim. The site elevation above Rogozna is 2.5 m. The terrace surface is not subject to inundation, yet the Rogozna flood stage rises a bit higher than the site cultural layer. Therefore human occupation of the site in the Late Pleistocene may serve a marker of ancient flood stages. Avdeevo had been extensively studied in terms of geochronology. To estimate the time interval of human occupation, we used the collection of radiocarbon dates published by Sulerzhitsky (2004) and constructed the summed probability function (Fig. 11B), which is discussed below.

4.2. Seim valley at Lgov

Hand coring profile across the large palaeomeander at the Kudintsevo village was produced by Panin et al. (2001) and then updated by Borisova et al. (2006). In both cases geochronology was based on few 14C dates. We have upgraded this profile using new results of mechanical coring to bedrock and OSL dating.

DGPS topographic profile crosses the 5.5-km wide floodplain segment formed by palaeochannel and neck of large meander and an isolated remnant of a 9-m terrace, probably lowered by aeolian action (Fig. 12A). As the top of the terrace remnant is occupied by a village with no allowance to drill, we tried to obtain the terrace alluvia for dating from the base of the remnant. The core S-H passed 2.7 m of yellow well-sorted fine sand interpreted as aeolian deposits blown out from the nearby terrace remnant. Below lied brownish loamy sand that was considered as channel alluvium, probably belonging to the terrace remnant. It was sampled from depth 5.5 m, but the obtained OSL date 19.5±1.0 ka is too young for Terrace T1. Therefore the dated unit was considered as Terrace T0 buried under thick aeolian apron.

Meander neck is composed of ridge-and-swale successions indicating lateral growth of palaeomeander. Under Holocene peat cover, swales are filled with 2-3 m thick loam deposits overlying yellowish fine sands that are considered to be channel alluvium. At levees, these sands lie immediately under the peat. Channel sands at palaeomeander neck were OSL dated to 17.1±1.2 ka and 16.4±1.2 ka. Palaeochannel is filled with heavy loams containing rare lenses of fine sand that were accumulated in an oxbow lake left after the palaeomeander abandonement. Base of palaeochannel infill was radiocarbon dated to 13800±85 BP (16690±160 cal BP) and 12630±70 BP (14980±160 cal BP). The palaeochannel surface has unusual topography; unlike in other filled palaeochannels usually characterized by smooth surface, the 2 m high and 200 m wide steep-sided ridge is found here composed of the heavy loam similar to that filling the palaeochannel (Fig. 12A). The ridge follows for a distance of 4 km along the palaeochannel axis and finally merges with the palaeochannel surface (Fig. 2A). The top of the ridge rises above the highest stage of present-day floods. To assess the age of this form, we made a core in its top and sampled at depth 2.5 m. Th sample gave OSL age of 14.5 ± 0.8 ka.

The Seim floodplain is crossed by River Prutische, a small right-hand tributary of Seim with catchment area around 600 km². Both in the lower course of the Prutische valley and within the Seim floodplain, Prutische makes large palaeomeanders exceeding greatly the dimensions of the present-day river (Fig. 2A). Three of these palaeochannels were cored and radiocarbon dated (Fig. 12A). All cores revealed a three-member composition: peat (1.2-1.6 m) underlain by heavy loams accumulated in residual oxbow lakes and channel alluvium composed of fine sand at the base. In cores S-11 and PR-15 the base of oxbow loams was dated to 14200±160 and 13830±110 cal years BP, respectively, which gives the younger limit of paleochannel abandonement. Top of channel sands lies 2-3 m below river level (102 m below its present-day bottom). In core PR-17 oxbow loams gave an older date of 16280±140 and their contact with channel sands is much lower: 5 m below river, or 4 m below the present-day Prutische river bottom.

7-m terrace of Seim was studied in the exposure at the right river bank near the Serebryanniy village – Section S-S (Fig. 2A, 12B). The section was subdivided into five units.

Unit S-S-1, 0.0-1.6 m: whitish well-sorted fine sand with thin planar lamination.

Unit S-S-2, 1.6-2.1 m: light silty loam, unevenly colored (darker spots on light-brown background) due to bioturbation; a number of molehills (krotovinas) descend to the underlying layer.

Unit S-S-3, 2.1-3.2 m: yellow medium sand with lenses of coarse sand and beds of brownish loamy sand.

Unit S-S-4, 3.2-6.2 m (visible): horizontal interbedding of yellowish fine sand and reddish-brownish loamy sand and sandy loam (1-2 cm beds).

The uppermost unit is considered as aeolian deposit, Unit 2 – overbank alluvium, lower units – channel alluvium. OSL dates exhibit inversion: Unit 3 gave OSL age of 20.9 ± 1.6 ka, Unit 4 – 17.4 ± 0.9 ka.

4.3. Svapa valley at Semenovka

We complemented the coring profile by Panin et al. (2001) and Borisova et al. (2006) with three cores down to bedrock (Fig. 13). Top of bedrock turned out to be smooth and lies at the same level under different morphological elements of the valley – 6-7 m below the Svapa River level. The valley has a complicated morphology with four major elements: Terrace T1, Terrace T0, aeolian massifs and large palaeochannels.

Terrace T1 9-10 m high bears numerous closed depressions 1-2 m deep probably of thermokarstic origin. Terrace is covered by 2 m thick loess-like silts underlain by more or less loamy fine sands, that were considered as alluvia of channel facies. It was sampled at depth 6.0 m and OSL dated to 46.9 ± 3.0 ka. Large palaeochannel 350 m wide, which is 7-10 times that of the present-day Svapa River (*cf.* Table 1) is filled with heavy loams with lenses of fine sand accumulated in a residual oxbow lake after the abandonement of the palaeochannel. The base and the top of this infill were radiocarbon dated to 14030±70 BP (17040±140 cal BP) and 11755±80 BP (13590±90 cal BP) respectively. If compared with the Seim valley, both the present-day peat-covered surface of the palaeochannel and the base of its infill are located relatively high above the Svapa River: 4.5 m and -1 – -2 m respectively (*cf.* with the Kudintsevo palaeomeander of Seim: +1.5 m and -6 m).

The other, smaller palaeochannel is found in the right part of the profile at the boundary of a large aeolian massif (Fig. 13). The two palaeochannels are separated by a 200 m wide steep-sided elongated remnant of a terrace, which narrow flat top rises at 9.5 m above River Svapa, i.e. to the same level as Terrace T1 at the left side. Cores SV-6a and SV-5 at the top and slope of the terrace passed a 1.5-2 m thick cover of light-brown light loam with lenses of fine sand interpreted as overbank alluvia. Below lies yellow-brownish fine to medium sand with rare lenses of loamy medium sand considered as channel alluvia. This unit was sampled at depth 5.7 m in core SV-6a and OSL dated to 18.3 ± 1.0 ka.

5. Discussion

The focus of this study is to establish the Late Pleistocene incision/aggradation rhythms, their amplitude and geochronology. To reach this aim, we have to synthesize data from all studied sections, correlate alluvial units between them and assess the age of different geomorphic members of the valleys. The next step would be to estimate vertical position of the river corresponding to different geomorphic levels, which will give the ground for the assessment of the magnitude of incision/aggradation events.

5.1. Correlation of sections and establishment of terrace age

5.1.1. Kurchatov reach, section ML1

Lower alluvial units of Section ML1 demonstrate that the valley was aggrading during the period ~80-50 ka BP (Fig. 8A), though nor the beginning neither the end of this aggradation is not recorded in the section. Given the rather deep position of channel alluvia dated to 50 ka BP, one can suppose that aggradation had continued after this date. However it could not continue till the end of MIS 3 because Terrace T0 has already been forming around 30-35 ka BP as follows from dating of the lowermost unit in Section ML2 (Fig. 8B). Probably, time moment around 40 ka BP would be a reasonable estimate for the completion of valley aggradation and start of incision.

Field study of the section and analytical data allowed reassessing of the initial interpretation of the hollow in the upper part of the section that was first considered as river palaeochannel. First, it was found that the hollow is filled mostly by non-alluvial deposits, which is not common for palaeochannels. Second, all OSL dates from the hollow fill refer to MIS 2. Therefore these sediments could not be accumulated in a palaeochannel because Terrace T1 had been abandoned already in the end of MIS 3. Consistent with all data is interpretation of the hollow as an erosion form – gully that dissected the edge of the terrace. OSL ages exhibit the inversion: Units ML1-4 and ML1-2 were dated at around 16-18 ka BP, while the uppermost Unit ML1-1 was dated to around 22 ka. We have two arguments in favor of the reliability of the latter date. First, the aeolian cover was followed over the terrace slope to Terrace T0 where it was dated to 19 ka BP. This date fits well to other data from sections ML2 and ML3 (see discussion in paragraph 5.1.2). Second, the hollow fill is subject to heavy cryogenic disturbance, which indirectly points at its deposition before LGM. Age underestimation by OSL dates around 16-18 ka may not be too much. Also, the erosion form could not have stayed stable for a long time within erodible sand deposits. Then, the gully formation and subsequent infilling may had occurred shortly before or in the very beginning of the LGM. Erosion dissection of the terrace may have not carry a climatic signal, rather it was facilitated by formation of terrace scarp due to preceding river incision.

5.1.2. Kurchatov reach, sections ML1 and ML 2

Sections ML2 and ML3 were correlated on the basis of geochronological data, results of pollen analysis and geomorphological implications. The latter was grounded on similar elevation of both sedimentary bodies and their close location in the valley, so that their sedimentation histories must have been coordinated in overbank deposition and response to river lateral migrations.

The lowermost lithostratigraphic unit in the exposed part of Section ML3 (Unit ML3-4) was not dated. However the composition of pollen assemblages in this unit (PZ ML3-1) is close to the sample at the depth 520 cm in section ML2, 10 cm below the OSL date 33.1±3.0 ka BP. Pollen records suggest an existence of boreal vegetation with coniferous forests dominated by spruce and pine with the admixture of birch. Evidently, Unit ML3-4 and Unit ML2-4 were formed during climatic warming in the end of the MIS 3 interstadial.

The middle parts of both sections (Unit ML2-3 – ML2-2 and Unit ML3-3) gave incomplete pollen data to be used in correlation. Their close age follows from OSL dates, all associated with MIS 2 (Fig. 7B, C). Sedimentation in both sections was not that synchronous. Channel alluvium of Unit ML2-3 had been deposited before the LGM. The lower boundary of the Unit represents the position of river bottom at that time (around 25 ka BP), which was almost 2 m above the level of the present-day river (3-4 m above its bottom). Obviously, alluvial sedimentation must have been occurring at the same elevation at the other river bank at that time unless there were no terraces composed by older deposits. However the corresponding Unit ML3-3 is much younger: three OSL dates point at its deposition immediately after the LGM. This is explained by destruction of the LGM and younger deposits by river lateral migration. The lower contact of Unit ML3-3 marks the position of river bottom around 20 ka BP: it was almost at the same level as 5 ka before, some 1.5 m above the present-day river. The lower half of Unit ML3-3 is channel alluvia deposited by extremely slow flow, the upper part – overbank alluvia deposited around 19 ka BP after the channel had moved away from the site.

At the left river bank, contemporaneous to the overbank alluvia in ML3-3 is the aeolian Unit ML2-2. Aeolian deposition in Section ML2 was facilitated by the proximity of Terrace T1, which served as a source of sand material. The sand blanket covering Terrace T1 was traced through its slope to Section ML2 where it dips under the younger overbank alluvia (Fig. 5). Such active aeolian action implies rather arid conditions that prevented stabilization of terrace sands by dense vegetation cover. Aridness at that time is supported by the abundance of carbonate concretions in Unit ML3-3 that indicate groundwater capillary rise and evaporation in soil profile.

Deposition of units 3 and 2 in both sections is divided by a discontinuity 4-5 ka long, which is indicated by a gap in absolute dates and by diagenetic changes in the top of Unit ML2-2. Long standing at the earth's surface is stressed by rise of pollen concentration at the top of aeolian sands of Unit ML2-2 (Fig. 9). To assess the chronology of this period, the AMS 14C dating of bulk organics was performed from the top of Unit ML2-2. Date around 10 ka obtained in section ML2 must be rejected as it contradicts to the older age of the overlying unit. The reason for age underestimation may lie in contamination of the top of sands by younger humus from the overlying humus-rich loams. The radiometric 14C date obtained from the same stratigraphic position in the terrace scarp (Section ML2a, see Fig. 8) is in consistency with the dates from the uppermost units in both sections. Probably this is due to the soil burial at larger depth and its better isolation from water filtration from the top of the section. This date suggests that soil burial occurred and accumulation of Unit ML2-1 started at or some after 14.8 cal ka BP.

Dates from the uppermost units (ML2-1, ML3-2 and ML3-1) make two groups. Date around 10 ka from Unit ML2-1 must be rejected: otherwise it impossible to explain the occurrence of sedimentation at the same elevation on one river bank (Section ML3) and its absence on another bank (Section ML2). The older dates around 14-15 ka are more consistent in that they are present in both sections and correspond well to the burial time of underlying soil. According to Chepalyga (2000), mollusk assemblages found in the overbank loams in Terrace T0 (Section ML2) indicate rather warm conditions and inhabit the river at present. This is not contradictory to dating these overbank loams by the Bolling-Allerod Interstadial.

Late Glacial age of units ML2-1 and ML3-2 is supported also by the results of pollen analysis. The maximum of *Picea* detected in PZ ML3-2 (Fig. 10) agrees well with the so called “lower maximum of spruce” – diagnostic feature for Allerød pollen spectra from the central part of the East European Plain (Khotinsky, Klimanov, 1997; Krementski et al., 2000). The pollen records reveal that the territory was covered by a complex vegetation of birch-pine and spruce-pine woodlands with meadows, shrubs thickets and plant communities on disturbed soils and sandy grounds. Similar palynological data were obtained for the fluvial deposits in the Svapa River valley about 80 km west from the study site (Borisova et al., 2006). According to these results the periglacial steppe plant communities also persisted in the region in favourable habitats. Pollen assemblages of studied sections include pollen of broadleaf deciduous taxa (*Quercus*, *Ulmus*, *Tilia* and *Corylus*). Though these trees are represented only by rare pollen grains, they might actually grew in the region, as confirmed by their frequent occurrence in pollen spectra of that time period in a number of sections in Eastern Europe (Feurdean et al., 2014). The presence of pollen of aquatic plants, relatively sensitive to heat supply, such as *Nymphaea alba* and *Typha latifolia* in pollen assemblages of study area (Borisova et al., 2006) suggests a rather warm climatic conditions.

Prominent feature of Section ML2 is that the overbank loams of Unit ML2-1 lay not only on the terrace surface but also wrap the buried terrace scarp and descend under the river level (Fig. 8). This indicates river incision before the deposition of Unit ML2-1 but after Unit ML2-2, i.e. between 15 and 19 ka BP. Later in the Holocene, the site avoided river lateral erosion and was buried by the Holocene overbank deposits. Such preservation is a rather rare stratigraphic feature in river valleys as terrace scars are almost always date from their subsequent undercutting rather than from the entrenchment event.

Pollen zones ML2-3, ML2-3 and ML3-3 in the upper 45 cm in both sections correspond to lithostratigraphic Unit ML3-1 and top of Unit ML2-1. The latter can not be separated from lithology, probably due to high degree of soil process reworking. The pollen assemblages are highly compressed and incomplete, but may be confidently said to represent typical Holocene successions of vegetation (Khotinsky, Klimanov, 1997; Novenko et al., 2015). The changes in pollen spectra suggest that birch woodlands were widespread in the initial phase of the Holocene, and then they were replaced by mixed broadleaf forests with pine. The obtained pollen data have shown that in the late Holocene the regional vegetation was strongly influenced by anthropogenic factors.

In contrast to Terrace T1, both sections in Terrace T0 make the impression of accumulation in rather warm environment because no permafrost features have been detected in any unit, including those subject to long exposition on the Earth's surface. Coincident with this is the fact that in both sections no alluvial sediments from the first half of MIS 2 have been found. This may arise from low fluvial activity in this time, both in terms of river lateral migrations and overbank deposition. Low elevation of floods in the first half of MIS 2 including the LGM is supported by data from the Avdeevo Upper Palaeolithic site (Fig. 9). There is strong evidence on multiple stationary occupation of the site by ancient humans on a year-round basis (Gvozdover, 1995; Sulerzhitsky, 2004), which implies that the site was not subject to flooding at that time. One of the evidences of non-seasonal human presence is the presence of dwellings and hearths. Dates on bone charcoal that was collected from hearths, dwellings and associated pits make three distinct peaks at around 27.2, 25.3 and 24.3 cal ka BP in the cumulative distribution of dates (Fig. 9B). Two small peaks produced by dates on bones collected outside any archaeological context evidence that humans visited the site also around 22.4 and 20.1 cal ka BP. Given that all samples were collected from a single cultural layer and were not separated by sterile alluvial deposits, the non-inundation period may have lasted at least between 20.0 and 27.5 cal ka BP. As was shown in Sections ML2 and ML3, period around 19 ka BP was also rather arid. Probably the aridness that marked the LGM did not finish with the termination of LGM around 20 ka but still continued to exist for one-two thousand years more.

5.1.3. Lgov reach of Seim and Svapa valley

These two sites are reasonable to be discussed jointly because they represent two confluent rivers that must have been interdependent in their response to environmental changes. Also, similarity may be expected with the Kurchatov reach located upstream in the Seim valley. Nevertheless, along with many common features, a number of specific details may be stressed for these sites, which contribute to our understanding of valley development.

Terrace T1 dated in the Svapa valley to around 47 ka BP (Fig. 13) is consistent in age with the same terrace at Kurchatov. What differs is that in Svapa this terrace is loess-covered and abundant with relic thermokarstic depressions unlike the terrace at Kurchatov, which is topped by the LGM aeolian blanket. Causes if this difference are unclear; most probably they have local origin, which shows that type of terrace mantle and vegetation cover may not be a reliable indicator of terrace age. More resembling the Kurchatov terrace is Terrace T1 at Lgov, which is also covered by aeolian sands (Fig. 12A). Both at Lgov and in Svapa, this terrace is a bit lower than at Kurchatov: 8-9 m against >10 m. This may arise from widening of the Seim valley downstream: the wide a valley is, the lower is a flood stage. This is appropriate also for the Holocene and for present day: both the elevation of floodplain and level of modern floods are by ~1 m lower at Lgov than at Kurchatov.

Terrace T0 at the Lgov reach studied in Section S-S (Fig. 12B) gave the inversion in OSL dates. If the lower and younger date of 17 ka was accepted, it would mean that valley aggradation continued after 17 ka, which contradicts to all other data on significant incision after 178-189 ka BP. Also, given that the terrace is mantled by aeolian sands, one must admit the occurrence of active aeolian processes in the Late Glacial time. There is no other reasoning such a conclusion. In particularly, high activity of aeolian processes at Kurchatov was dated at 19-20 ka, i.e. around the LGM. According to this reasoning the upper and older date of 21 ka would be more appropriate. However the whole section does not leave an impression of being accumulated during the LGM: no cryogenic features were found; on the contrary, abundant root casts point at well-developed vegetation cover at the final stage of terrace accumulation before it had been mantled by aeolian sands. Probably, not only the younger, but also the older date underestimates the age. If the date was 3-4 ka older and referred to the pre-LGM time, the terrace would be quite similar to Terrace T0 in section ML2 at Kurchatov.

Quite different is terrace T0 in the Svapa valley. Unusual is its elevation, which is similar to that of Terrace T1. However its OSL age measured at ~18 ka is consistent with geomorphological reasoning that this terrace gave start to the incision in course of formation of large palaeochannels at both sides of the terrace remnant (Fig. 13). Possible reasons for such kind of morphology is discussed below. Probably this is the high position of terrace T0 that caused the adjacent large palaeochannel to insufficiently deepen relative to the modern river: base of palaeochannel fill is only 2 m below the modern river while in the Seim valley it reaches 6 m below the river. At the same time, the elevation range between the Terrace T0 top and the palaeochannel bed is similar in Seim and Svapa and constitutes 10-11 m, which implies also similar depth of incision.

Two phases of post-LGM fluvial activity related to increased water discharges may be deduced from the Seim and Svapa large palaeochannels that demonstrate the presence of two generations. The first generation showed in profiles in Figs 12A and 13 was developing between 18-17 ka BP, maybe till 16 ka BP as is shown by the abandoned large meander of River Prutische (see section PR-17 in Fig. 12B). The second phase has not been preserved as large palaeochannels of Seim and Svapa, most probably because they were not abandoned and has been preserved till nowadays as the bend of modern meandering belt (Fig. 2A). However there are two kinds of geomorphic evidence in favour of this phase of fluvial activity. First is large palaeomeanders of the Prutische River dated to around 14 ka BP (see sections S-11 and PR-15 in Figs. 2A and 12B). Second is large ridge in the abandoned palaeochannel of Seim dated to ~14.5 ka BP, which was formed most probably by flood waters that used the then not totally filled palaeochannel when moving over floodplain (Figs. 2A, 12A). The above dates are consistent with the dates on overbank loams at Kurchatov reach (Fig. 7B, C) considered in the above to be accumulated due to considerable rise of flood levels. Given the dating results, this phase of increased discharges and flood levels lasted during 15-13 ka BP.

5.2. Time and magnitude of incision/aggradation events

To assess the chronology and amplitude of incision and aggradation events we analyzed the elevation of alluvial units of different age on the basis of the dates in Tables 3 and 4. At first inappropriate dates were filtered out such as non-alluvial dates and dates that do not fit the stratigraphic context (see the above discussion). The rest 30 dates were plotted against elevation above modern rivers separately for different alluvial facies (Fig. 14). The thick dashed line stands for position of river channel bed. Alluvial bed at 25 ka BP was identified in section ML2 (1.6 m above the river), that of 20 ka BP – in section ML3 (1.5 m above the river) (Fig. 7B, C, Fig. 8). For other periods the line envelopes the cloud of points and approximates the position of the channel bed.

The above results can be summarized in the following succession of incision and aggradation phases.

1. *Incision preceding the aggradation of Terrace T1*. It has not been clearly dated, but definitely finalized before ~80 ka BP (the lower alluvial units in section ML1). Mikulian (Eemian; MIS 5e) age was reported from the alluvial base of this terrace both in the Seim valley (Chepalyga, 2000) and in other valleys in central Russian Plain (Gurtovaya, Faustova, 1977; Ivanova, Tiurina, 1979; Rychagov, Antonov, 1996). Then the incision must have had pre-Mikulinina, probably late MIS 6 age. Probably it was this phase that the incision reached the bedrock that lies at large depth beneath the modern rivers. In the Kurchatov reach, bedrock under Terrace T0 and the floodplain was reached at 7-11 m below river (Fig. 5). Even deeper is the bedrock surface in the Lgov reach – 15-16 m (Fig. 12A). The least deep is bedrock in the Svapa valley – 6-7 m below the river (Table 5; Figs 5, 12, 13). Horizontal surface of bedrock under different elements of the Svapa valley such as Terraces T1 and T0 and floodplain (Fig. 13) indicates that bedrock was reached and leveled by river lateral migration already before the aggradation of Terrace T1, most probably during the pre-Mikulinian incision phase.

2. *Aggradation of Terrace T1*: started in MIS 5, before ~80 ka BP, and finished not earlier than in the middle of MIS 3, which is supported by OSL dates around 47-50 ka BP from Terrace T1 in the Kurchatov reach of Seim and in Svapa valley (Figs 5, 13).

3. *Incision into Terrace T1* may be placed in the interval between 35 – 45 ka BP based on the above mentioned dates from Terrace T1 and dates around 32-33 ka BP from the base of terrace T0 at Kurchatov reach (Fig. 6). Minimum depth of incision is exhibited by Unit ML2-4 indicative of lateral movement of the river channel at around 32-33 ka BP at an elevation similar to that of the present-day river (Fig. 6). The maximum incision could have been deeper and have occurred at some earlier time: top of channel alluvium in section ML3 was reached at 3 m below the river.

4. *Aggradation of Terrace T0:* started already in the end of MIS 3 and continued in MIS 2. The major part of aggradation had occurred probably in the first half of MIS 2 prior to the LGM as follows from sections ML2 and ML3 showing similar position of channel alluvia in the interval 25-20 ka BP, i.e. right before and right after the LGM. In this sections Terrace T0 demonstrates rather small thickness of MIS 2 alluvium – only 3-4 m of 6-7 m of the terrace height above the river. The base of the terrace is composed of alluvium from the late MIS 3. In other sections, particularly in the Lgov reach of Seim, channel alluvium from around the end of the LGM was found also at the level of the modern river. An exception is presented by the Svapa valley where Terrace T0 has the same or similar elevation as Terrace T1 (Fig. 13). Two interpretations may be suggested. First, it could have been shaped in such a way if there were no incision in the end of MIS 3. This explanation looks improbable because of occurrence of deep incision in the Seim valley in late MIS 3, which could not have stayed without causing the response in tributary valleys such as Svapa. Second explanation may stand in thick aggradation in early MIS 2 due to some local causes such as excessive input of sand material. Evidence for such input may lie in vast aeolian sand massifs common for this reach of the valley (Fig. 2A), but the origin of this sand material is unclear yet.

5. *Incision into Terrace T0*. Start of this incision at about 18 ka BP is supported by a number of dates from inner parts of large meanders in the Lgov reach of Seim and in Svapa (Figs 12, 13), as well as by temporary interruption of sedimentation after ~18 ka BP on Terrace T0 in Section ML3 (Fig. 7C). Valley deepening reached the maximum depth very shortly: already around 17 ka BP the channel bed in the thalweg of the abandoned large meander at Lgov stood at 6 m below the modern river (Fig. 12A). Channel position below the modern river at ~14 ka BP is evidence by overbank loams that mantle terrace scarp in sections ML2 – ML2a (Fig. 8).

6. *Aggradation to the present-day position*. It had occurred in most part at the Younger Dryas – Holocene transition because small meandering palaeochannels with 14C ages 9-10 uncal ka BP already correspond to the position of present-day river channel.

The described succession of events and produced valley morphology is consistent to much data reported from Central and Western Europe, though some dissimilarity can also be found.

Antoine et al. (2007) proposed that the main incision leading to terrace formation occurred during early glacial phases, i.e. can be attributed to the transition between interglacial and glacial. This is not the case in the Seim River catchment where incision and terrace formation took place in the second half of the last glacial epoch. In fact, the late MIS 3 incision may be regarded as the major valley deepening phase in the Late Pleistocene as it was during this phase that the scarp between the wide spread Terrace T1 and valley bottom was formed. The last Late Pleistocene incision phase in the Late Glacial was almost compensated by the subsequent aggradation at the transition to the Holocene, so that the LGM Terrace T0 differs very slightly in elevation from the Lateglacial – Holocene floodplain and was even subject to inundation in some periods in the Holocene. In the upper Thames, no incision was detected at the beginning of the Early Glacial (MIS 5 – MIS 4 transition) and aggradation tendencies continued throughout the Pleniglacial till the LGM and were followed by incision only at the start of the Late Glacial warming (Lewis et al., 2001). In River Exe in South West England two marked incision events were revealed in the Late Pleistocene, of which the first had occurred between 60 ka BP and 30 ka BP (MIS 3) and the second dated to around the LGM (Brown et al., 2010).

Rose and Meng (1999) and Rose et al. (1999) found that in Mallorca, Spain, the last 140 ka, episodes of fluvial aggradation occurred during MIS 4 and late MIS 3 through into very early IOS 1 with greatest levels of fluvial aggradation occurring in MIS 2. In a small river valley in Slovakia, Nowaczinski et al. (2015) found several incision episodes below the modern rive level around 35-42 cal ka BP and another incision after LGM between 14-17 cal ka BP. Unlike in other central and East-European valleys, the post-LGM incision was not followed by transformation of river style from braided to meandering. Multiple incision/aggradation events during the last glacial epoch were found in Tagus valley, Portugal, with aggradation events being coincident with cold episodes and incision events characteristic to climate warmings (van den Schriek et al., 2007). In River Aguas in the southern Iberian Peninsula, valley aggradation since Eemian till MIS 2 was found followed by deep incision in MIS 2 before LGM and two other incision events in the end of the Late Glacial (Schulte et al., 2008). This succession is quite similar to what has been found in Seim.

The proximity of elevation of Pleniglacial and Holocene alluvial terraces characteristic for River Seim has been reported also from a number of valleys in Central Europe, such as the Ner River in Central Poland where the base of the terrace date to around LGM lies noticeably lower the Holocene channel alluvia and the surface of the same terrace lies at the same level as the alluvial plain accreted in the Holocene (Kittel et al., 2016). Like in the Kurchatov reach of Seim, in the Upper Danube River in Bavaria aggradation in MIS 2 had been completed before the LGM and since 23-24 ka BP gradual incision had been occurring that accelerated considerably after 16 ka BP (Heine, 1999). Schirmer et al. (2005) recognize in the Upper Danube and over the Rhine catchment (Rivers Main, Isar, Weser, Upper and Lower Rhine) three Late Weichselian terraces which aggradation was interrupted by incision events at 23-24, 13-14.5 (Late Glacial warming) and 11.6 cal ka BP (the Pleistocene/Holocene boundary). In the Seim River aggradation of Terrace T0 also seems to have finished before the LGM and since ~25 ka BP lateral channel migration prevailed. However incision startet in Seim a bit afdter the LGM, around 18 ka BP.

Considerable increase of runoff and formation of large meanders during the Bølling-Allerød Interstadial was found in upper Thames (Lewis et al., 2001), in most Polish rivers (Schumanski, 1983; Vandenberghe et al., 1994), in Rivers Tisza (Kasse et al., 2010) and Teleorman (Howard et al., 2004) in the lower Danube catchment. The most precise dating of the first post-LGM incision was achieved in the Jeetzel River (lower Elbe valley) where traces of the main Lateglacial temperature rise were detected in palaeochannel fills associated with incision into the Pleniglacial braidplain, which means that incision had taken place a little before 14.6-14.7 ka cal BP.

On the other hand, in many European valleys two post-LGM impulses of incision were found both being associated with transformation of braided rivers into single-thread, usually meandering channels. The first post-LGM incision phase is dated to the onset of the Late Glacial warming between 14-15 cal ka BP, the second one occurred at the onset of the Holocene. Both incision phases were found in northern France (Antoine et al., 2003), in lower Maas (Vandenberghe et al., 1994), in Upper (Erkens et al., 2009) and Lower Rhine (Erkens et al., 2011), though according to van Balen et al. (2010), the first post-LGM incision in the lower Rhine-Meuse fluvial system had been delayed till the start of the Younger Dryas, in northern Germany (Turner et al., 2013), in most Polish rivers (Vandenberghe et al., 1994; Petera-Zganiacz et al., 2015; Starkel et al., 2015). In the Upper Rhine no indication of a Younger Dryas braid plain was found, and the incision at the end of the Late Glacial was succeeded into the Holocene and resulted in formation of the Holocene terrace staircase, most probably in response to local tectonic movements (Erkens et al., 2009). Slight incision through the Holocene was reconstructed also in major tributaries of Rhine (Schirmer et al., 2005). Also, in the Upper Danube valley, up to eight Holocene terraces were found at the Lech confluence confirming continuing incision tendencies (Schielein et al., 2011), and Mid- to Late Holocene entrenchment was reported from the Lower Danube tributaries (Howard et a;., 2004), though prevalence of lateral sediment reworking with no significant incision in the Holocene was reported from a number Danube locations in Bavaria (Heine, 1999). No incision at the transition to the Holocene was found in Seim, by contrast, this transition was marked by river channel aggradation after considerable incision in the Late Glacial.

Much influence was exerted on European valleys by non-climatic factors. In the lower Rhine-Meuse fluvial system, incision was found at the MIS 3 to MIS 2 climatic transition (van Balen et al., 2010). Given that this incision is not reproduced by a numerical landscape evolution model, the authors refer it rather to other (glacio-isostatic?) than climatic reasons. Influence of glacioisostatic effects was probably responsible for development of the East German rivers that demonstrated continuous aggradation since late MIS 5 till MIS 2 (Mol, 1997; Kasse et al., 2003) interrupted by probably lateral erosion with no incision; probably dynamics of these rivers was influenced by glacioisostatic movements. In southern and central Poland, river aggradation was found to dominate during the cooler phases of the Vistulian and during the Interpleniglacial while two major incision phases bracketed LGM: they were dated to 25–20 ka BP and to the Upper Pleniglacial–Lateglacial transition at 15–13 ka BP (Starkel et al., 2007; Gębica et al., 2015). The authors stress that the second incision phase was not driven by climatic factors alone but coincided with the rapid downcutting in the lower course of the main Vistula valley after removal of the blockage by the Scandinavian ice sheet.

In the Russian Plain, the territories to the north and north-west from the study area closer to the former margin of the Scandinavian ice sheet, such as the upper Dnieper (Kalicki, Sanko, 1998; Panin et al., 2015), Western Dvina (San'ko, 1987; Kalicki et al., 1997), Neman (Voznyachuk, Valchik, 1978), upper Volga (Obedientova, 1977), Northern Dvina (Lyså et al., 2014; Zaretskaya et al., 2014) were subject to glacial damming and glacio-isostatic crustal adjustment, which hindered the climate forcing of river valley development. Lower reaches of major river basins that drain the Southern Russian Plain – lower Dniester, lower Dnieper, lower Don and lower Volga – experienced the influence of base level change due to the high-amplitude fluctuations of the Black Sea and Caspian Sea levels in the Late Pleistocene (Goretskiy, 1970; Matoshko et al., 2002, 2004; Matoshko, 2004). Similar complications of climate forcing in river development existed in the Danube and Vistula basins in Central Europe (Starkel et al., 2015). In the central regions of the Russian Plain, no allogenic forces but climate change governed river valley development. Based on similarity of valley morphology and the sporadic existing data on geochronology of floodplain and terrace units (Panin et al., 2011, 2013; Panin, Matlakhova, 2013), we believe that the established rhythms were manifested over a vast territory of Central Russian Plain comprising the middle Dnieper, Oka, Don and middle Volga basins.

5.3. Drivers of river incision and aggradation

A number of authors (Vandenberghe, 1995, 2003; Kasse et al., 2003) attribute the Lateglacial river incision in western and central Europe to the reduction of slope erosion and sediment supply to rivers due to the increase in vegetation cover. However immediate drop of river sediment load in response to decrease of denudation rates over catchment area seems hardly possible due to inertial effects caused by considerable residence time of sediment moving through a fluvial system. This is well illustrated by findings of van Balen et al. (2010) in lower Rhine-Meuse river system where numerical model of landscape evolution, in accordance with geological record, showed neither incision nor aggradation in the beginning of the Bølling–Allerød interstadial despite fast rise of vegetation cover density and reduction of sediment input. This effect is explained by high sediment storage accumulated within valley bottoms previously and available for fluvial transport. On the other hand, there are a number of studies suggesting that both the Early Lateglacial and Holocene onset incision events occurred some before the main temperature ameliorations and modifications of the vegetation cover (Antoine et al., 2003; Turner et al., 2013).

From our point of view, in the case of the Early Late Glacial, the lag between river incision and major thermal rise can be explained by admitting that the river incision was initiated by considerable rise of water runoff, which is evidenced by formation of large meanders with both wavelengths and channel width much greater than that of the Holocene river channels. Also, new data are emerging that consider change of river hydrological regime during the YD from snowmelt to rain-dominated (Andres et al., 2001) and propose clustering of flash flood as the cause for river incision at the decline of YD (Petera-Zganiacz et al., 2015). On the other hand, rivers in Southern England demonstrate occurrence of strong incision at the start of YD, which is suggested to have resulted from increased nival peak discharge under still resistant river banks densely covered by vegetation during the preceding Windermere (Bølling-Allerød) Interstadial that required time to respond to the changing climate (Gao et al., 2007).

Bridgland and Westaway (2008) conclude from their worldwide review of river terrace development that temperature fluctuations seem to be the key forcing factor of incision/aggradation rhytms with the role of varying humidity yet to be demonstrated. In our study, we have strong evidence that the post-LGM incision/ aggradation rhythm was caused by the river long profile adjustment to changes in runoff. Two kinds of arguments may be proposed based on the geological imprint of changes in flood levels and on changing morphology of river channel.

The elevation range of alluvium of similar age indicates the amplitude of river level changes, i.e. the flood magnitude (Fig. 14). Around the LGM, elevation range of alluvial sedimentation was at its minimum of 3-4 m only. This corresponds to a shallow braided channel where height of flood stages is greatly limited by large width of the valley floor. The post-LGM interval 18-12 ka BP is characterized by the highest range of elevations of the accumulated alluvium with channel fills lying below -4 m and overbank facies found at >5 m relative to the modern river, which gives the total range of at least 10 m. This means that the river bed stayed lower and floods rose higher than that of now. For example, around 14.5 ka BP, inundation of the large palaeochannel at Lgov led to accumulation of a loamy ridge with its top standing higher than the present-day maximum level of floods (Fig. 12A) while channel alluvium in large palaeochannels of the Seim tributary Prutische River active at the same time stands lower than that in the modern river. In the Kurchatov reach, the Late Glacial overbank alluvia dated to about 14 ka BP mantles the terrace scarp may be tracked down under water showing the position of the then active channel below the modern river (Fig. 8). At the same time, accumulation of overbank alluvia was found at the other river bank at elevation well above the present-day floods (Fig. 7C). In the Holocene the river bed aggraded and floods became lower: the range of alluvial sedimentation decreased to 6-7 m. Aggradation had occurred already by the very beginning of the Holocene: small meandering palaeochannels with reported 14C dates around 9-10 ka BP (uncal) occupy the same position as the present-day river channel (Borisova et al., 2006). The Holocene highest flood levels were observed probably around 2.5 ka BP (shown as a narrow spike in Fig. 14). This is indicated by accumulation of overbank alluvia on the surface of the 6 m terrace T0 in section ML3 dated by the finds of Early Iron Age ceramics (see section 4.1.3, Fig. 7C)

Distinct geomorphological evidence of considerable increase of channel-forming discharges in the Lateglacial is provided by the post-LGM single-thread palaeochannels, which width is by an order of magnitude larger than that of the Holocene palaeochannels. As was shown at the Lgov reach in Seim and Semionovka reach in Svapa valleys, starting from around 18 ka BP large meanders began to form and at the same time river incision had started. Not only palaeochannel size, but also pollen data indicate that water runoff increased considerably: Borisova et al. (2006) estimated average annual runoff depth in the Seim catchment about 15 ka BP at 300 mm, which is 2.5 times greater than the present-day value. This is in line with findings by Magny (2003) who points at increase of annual precipitation at the beginning of the

Lateglacial combined with lake-level lowering in western Europe. Magny argues that such kind of response may have been associated to climate seasonality characterised by moist winters and dry summers: lake-level lowering could have been provoked by summer dryness, while river incision may have resulted from higher discharge in winter. The continental climate of Eastern Europe with long cold winters, which duration and coldness were even higher in the post-LGM epoch than at presents (Borisova et al., 2006), makes impossible the occurrence of high runoff events in winter. Therefore we suggest that the high channel-forming discharges that produced the huge palaeomeanders of rivers Seim and Svapa occurred during spring snowmelt floods. Incision was caused by the adjustment of river long profiles to rise of runoff.

Less indication is for the factors of the late MIS 3 incision phase because of limited geomorphological and sedimentological record. However, no arguments for other than climate change driving forces exist. Also, some evidence has previously been reported from adjacent basins of occurrence of large palaeochannels that indicate high river runoff. Large meander dated to around 38 ka BP has been reported from the adjacent Don River basin (Panin et al., 2011). Active growth of incised large meander was dated to around 35 ka BP in the Moskva River north from the study site (Panin, Matlakhova, 2013). These examples make round for us to suppose that it was rise of river runoff similar to that happened in the Late Glacial that caused River Seim incision in the late MIS 3.

6. Conclusion

Two full incision/aggradation rhythms were detected in River Seim and its tributaries after the long aggradation phase that lasted probably since MIS 5e and was finalized by the middle of MIS 4. Deep incision that led to formation of Terrace T1 had occurred around 40 ka BP. Moderate aggradation started already in the end of MIS 3 and continued in the first half of MIS 2. Since ~25 ka and through the LGM lateral migrations occurred of a shallow braided channel located few meters above the present-day river. Terrace T0 was created that consists of late MIS 3 alluvium at the base overlain by MIS 2 channel and overbank alluvium. Incision into terrace T0 started ~18 ka BP and proceeded below the modern river. Multiple-thread channel concentrated in a single flow, which at some places formed large meanders. In the period 15-13 ka BP, high floods that rose above the present-day floods left large levees and overbank loams on Terrace T0. At the same time river channel stood few meters below the modern river. At the transition to the Holocene, moderate aggradation took place so that in the Early Holocene both vertical position and channel pattern of rivers were already similar to the present-day river and have not changed noticeably in the Holocene.

The result of river development in the late Pleistocene that included two incision and three aggradation phases was the formation of three terraces, of which the two lower terraces – Terrace T0 and Floodplain – have very small difference in elevation and often merge. The main scarp in the bottom of the valley between Terrace T1 and Terrace T0/Floodplain was formed in the late MIS 4 incision phase.

Incision phase in the end of MIS 2 (after the LGM) was undoubtedly governed by climatically forced large increase of water runoff. This phase consisted probably of two sub-phases at 18-16 and 15-13 ka BP separated by a relatively sub-phase of lower runoff. According to indirect reasoning, the late MIS 4 incision phase was also induced by considerable increase of water discharges.

Similarity of valley morphology and the absence of other allogenic forces but climate change, such as base level change or glacial damming and crustal movements caused by glacial loading, makes it most probably that the established rhythms were characteristic for river valleys over the major basins of the Central Russian Plain – the whole middle Dnieper, Oka, Don and middle Volga basins.

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